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DOWNVALLEY GRADIENTS IN FLOW PATTERNS, SEDIMENT TRANSPORT AND CHANNEL MORPHOLOGY IN A SMALL MACROTIDAL ESTUARY: DIPPER HARBOUR CREEK, NEW BRUNSWICK, CANADA

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ABSTRACT

Dipper Harbour Creek's lower reaches run through a narrow salt marsh on the Bay of Fundy, New Brunswick, Canada. This 2 km long section of the creek constitutes an extreme example of a tide-dominated estuary exhibiting strong downvalley morphology and sedimentology gradients. Dipper Harbour Creek drains a basin of roughly 8·8 km^2, but except during the spring snowmelt freshet, tidal flow so overshadows freshwater flow within the salt marsh reach that the system essentially functions as a tidal creek. To identify and explain the main geomorphic processes controlling the creek system, records were collected in summer 1993 of tidal stage and velocity fluctuations, sand dune migration rates, bed material composition, channel cross-sectional geometry and channel sinuosity. Bed materials become progressively finer upvalley, with deposits of medium to coarse sands concentrated in the highly sinuous central reach of the creek during the summer. Current velocities within the creek are strongly flood-dominant, featuring a consistent low-stage peak in flood velocity, a secondary high-stage flood surge, and a weaker ebb peak occurring around bankfull stage. Under summer low freshwater discharge conditions, the predominant direction of bed sand transport is upvalley. The spring freshet, however, causes a major downvalley shift of sand deposits, suggesting a seasonal cycling of medium to coarse sands within Dipper Harbour Creek.

KEY WORDS estuary; salt marsh; macrotidal; tidal current; velocity asymmetry; Bay of Fundy

INTRODUCTION

In their general facies model of estuaries, Dalrymple *et al.* (1992) point out typical downvalley patterns in the morphology and sedimentology of tide-dominated estuaries, including a 'straight–meandering–straight' pattern of channel alignment (their figure 7). These three reaches are said to be controlled, respectively, by fluvial, mixed and tidal processes. The two straighter reaches feature a single dominant direction of sediment transport while the central meandering reach has little net sediment motion and contains the estuary's finest bed material. This general zonation is supported, for example, by observations on the estuaries of the Salmon River (Dalrymple and Zaitlin, 1989) and the Ord River (Wright *et al.*, 1975). The present case study focuses on Dipper Harbour Creek, a small coastal New Brunswick stream whose lower reaches form a macrotidal estuary fringed by salt marsh. The tidal portion of Dipper Harbour Creek exhibits strong downvalley gradients in morphology and sedimentology, suggesting some interaction of tidal and fluvial forces. This paper examines these forces and their effects on channel morphology in the lower reaches of this system. In particular, it attempts to illustrate how the generalized dynamic balance described by Dalrymple *et al.* is realised in the extreme case of a small estuary where fluvial forces are only seasonally effective.

Several studies have been conducted on the morphology and sedimentology of macrotidal estuaries (e.g.

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Wright *et al.*, 1975; Ashley and Renwick, 1983; Woodroffe *et al.*, 1989), but the systems are generally much larger and of more consistent freshwater input than Dipper Harbour Creek. Except during the brief springtime snowmelt freshet, tidal flow so overshadows freshwater flow in Dipper Harbour Creek that for most of the year it essentially functions as a true tidal channel and can be considered analogous to a salt marsh tidal creek.

Pure tidal creeks are a common feature of many coasts, and a substantial body of literature exists concerning their arrangement and physical processes. Research in this area has variously dealt with the development of tidal creek drainage networks (Pestrong, 1965; Knighton *et al.*, 1991, 1992), suspended material exchange through the creeks (Boon, 1974; Settlemyre and Gardner, 1977; Ward, 1981; Reed, 1988), discharge asymmetry (Boon, 1975; Reed, 1987), hydraulic geometry (Myrick and Leopold, 1963), turbulence patterns (French and Clifford, 1992), point-bar sedimentology (Barwis, 1978) and the role of wetland vegetation in creek morphology (Garofalo, 1980).

A relatively small number of detailed examinations of tidal creek hydrodynamics exist (Bayliss-Smith *et al.*, 1979; Pethick, 1980; French and Stoddart, 1992) and they are especially pertinent to this study. Some published work on mudflat creeks (Adams *et al.*, 1990; Wells *et al.*, 1990) and shallow estuaries (Shetye and Gouveia, 1992) also provides relevant, detailed information on tidal flow in small channels. Clearly, tidal hydrodynamics are transformed in complex ways by the particular morphology of each coastal drainage system and its sedimentary fill. While pure tidal creek systems tend to be ebb-dominant (Petsrong, 1965; Bayliss-Smith *et al.*, 1978; Reed *et al.*, 1985; French and Stoddart, 1992), shallow macrotidal estuaries are often flood-dominant (Dalrymple and Zaitlin, 1989; Wright *et al.*, 1975).

The present study will describe the flow patterns, sedimentology and morphology of Dipper Harbour Creek. Results will be presented in turn on channel form, bed material composition, tidal flow patterns and bed sediment transport. Two main questions will be examined in this light. Firstly, given the strongly tidal regime of Dipper Harbour Creek during the low runoff summer season, how do the hydrodynamics of this estuary compare to those of the true tidal creek systems described in the literature? Secondly, does the combination of very strong tides and highly seasonal fluvial energy seen in Dipper Harbour Creek modify the general sedimentological model for macrotidal estuaries described by Dalrymple *et al.* (1992)? This second question refers particularly to the sedimentary dynamics of the central reach where fluvial and tidal forces achieve some average balance.

STUDY AREA

Dipper Harbour creek, located 35 km southwest of Saint John, New Brunswick, empties into Dipper Harbour on the Bay of Fundy's north shore (Figure 1). Its lowest reaches, subject to the bay's strong tides, run through a narrow 1·7 km long salt marsh whose major plant species include *Spartina alterniflora*, *Spartina patens*, *Plantago maritima*, *Triglochin elata*, *Juncus gerardi* and *Salicornia europia*. A decimetre-scale topographic gradient separates the frequently flooded, *S. alterniflora*-dominated low marsh from the less frequently flooded, *S. patens*-dominated high marsh. In a few locations near the mouth, the marsh surface slopes gently toward the channel, but generally the creek banks are very abrupt. Occasional slumped blocks of low marsh are found in the channel. The creek drains a forested 8·8 km^2 basin which receives net rainfall of about 900 mm per year, producing a mean annual discharge of roughly 0·25 $m^3\ s^{-1}$. In May 1994, a month after the annual spring meltwater flood peak, a freshwater discharge of 0·85 $m^3\ s^{-1}$ was observed in the upper creek after a strong rain event. At the time, the stream entered the marsh carrying a suspended sediment load of roughly 70 $mg\ l^{-1}$, half of which was medium sand (particle diameter greater than 0·212 mm). Based on survey data and May 1994 discharge, the creek's mean annual freshwater flood discharge from the upland is estimated at between 2 and 3 $m^3\ s^{-1}$. For most of the year, however, freshwater inputs are under 0·1 $m^3\ s^{-1}$ and lower Dipper Harbour Creek is functionally a tidal creek.

The Saint John region of the Bay of Fundy has large tidal ranges, reaching 8·5 m on some spring tides. The mouth of Dipper Harbour Creek, perched at the head of the bay of the same name, is exposed to only the top portion of each semi-diurnal Fundy tidal cycle, but spring tide amplitudes of over 3 m are still observed at the creek mouth. Once the creek has drained after high tide, a slow ebbward trickle continues until the onset of the next flood, for a total ebb flow duration of roughly 10 h, compared to only 2–3 h of flood. Based on

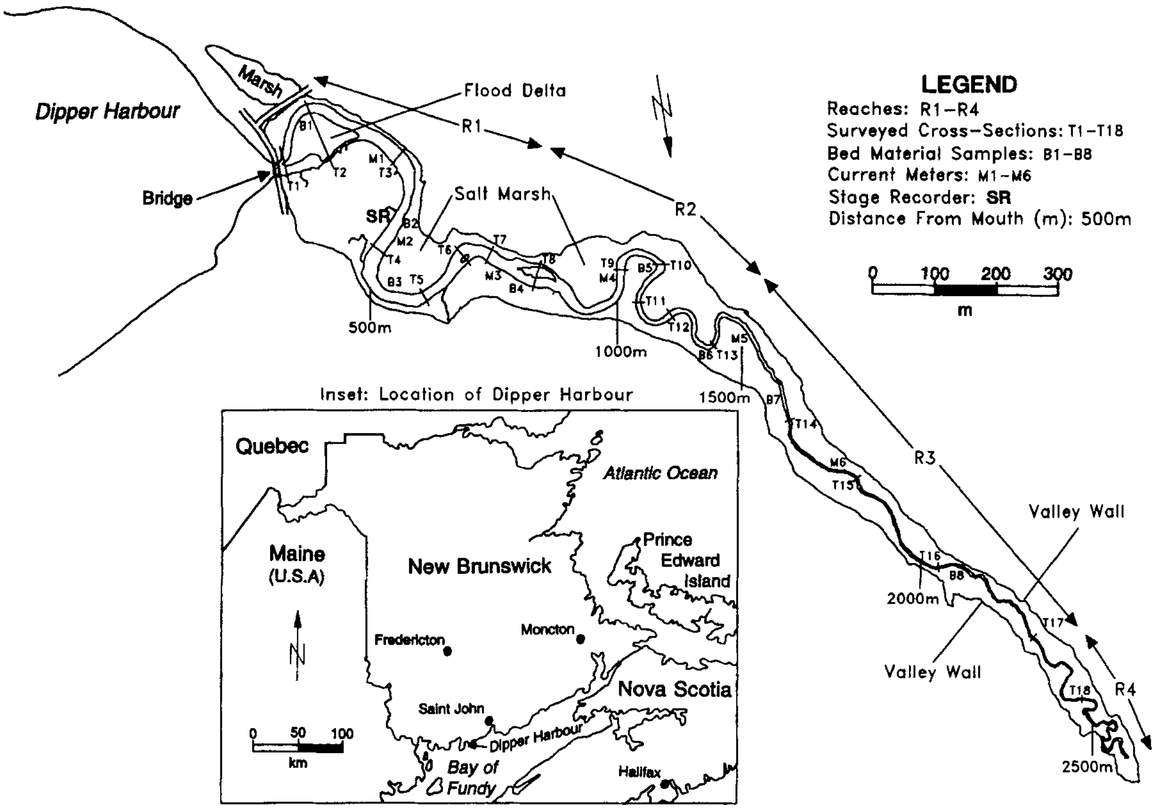

Figure 1. Location map showing study reaches, channel transects, bed material sites, current meter and stage recorder positions. Inset shows location of Dipper Harbour Creek in New Brunswick, Canada

detailed tidal prism volume estimates (Ayles, 1993), a large spring tide brings about 260 000 m^3 of sea water into the marsh–creek system, yielding a tidal discharge of 25–35 $m^3 s^{-1}$ averaged over the flood limb, which is over ten times the typical spring freshet discharge and over 100 times the mean annual freshwater discharge. Dipper Harbour Creek would lie at the very extreme of tide-dominance in the estuarine classification of Dalrymple *et al.* (1992).

Dipper Harbour Bay is a mudflat at low tide, with coarse gravels and sands found only in the creek's outflow channel, and the creek and harbour are separated by a road bridge built upon a causeway of crushed rocks and gravel. Historic maps show that the causeway and road were built early this century upon a Holocene spit which blocked the mouth of the valley, cutting off wave energy from the creek and marsh. Sea water can only enter the marsh–creek system by funnelling through the 11 m wide bridge opening and this flow constriction has created a gravel flood delta just upvalley of the bridge. Within the salt marsh, there is a single dominant channel, through many minor gullies (less than 1 m wide) and collapsed groundwater channels fork out into the marsh. A few small tributary freshwater brooks also enter the marsh, and a short tidal creek branches off the flood delta into a small marsh to the south.

METHODS

Channel width and longitudinal profile data were surveyed in the field. Eighteen cross-sections (Figure 1) were surveyed at intervals across the channel and its adjacent low marsh along the creek length. Bed material composition was obtained by volumetric sampling of the bed. Samples were taken near the channel centre in locations (Figure 1) representative of the overall textural quality of each subject reach. Several litres of substrate, extracted from a maximum depth of approximately 15 cm, were taken from eight sites along the creek. No particles with a *B*-axis diameter of greater than 10 cm were found in any of the samples.

Current velocity data were obtained using two Marsh–McBirney 511M electromagnetic current meters (sensor head diameter 4 cm). The sensor electronics and datalogger for the first of these meters were stored in a watertight aluminium cylinder, allowing the entire apparatus to be submerged in the channel. The second instrument's electronics were sheltered from the tide on a platform on the marsh surface adjacent to the channel. In all deployments, sensors were mounted 15 cm above the creek bed near to the centre of relatively straight reaches to optimize the gathering of both flood and ebb data, and 5 min average velocities were recorded. Tidal velocity data were gathered at six stations (Figure 1) for a total of 17 deployments between 17 July and 22 August 1994, over tides with amplitudes between approximately 4 and 138 cm over bankfull. The present paper will concentrate on the results of four deployments at Positions M3, M4 and M6, which provided uninterrupted flow sensing, including simultaneous recording at Positions M3 and M4 between 20 and 22 August. As will be shown later, peak flows in Dipper Harbour Creek occur early in the flood, and at Positions M1, M2 and M5 the sensor emerged at low tide, resulting in the loss of these crucial low-stage data.

Water level changes in the creek were charted using a float stage recorder installed in a stilling well in the lower marsh (Figure 1). Tidal stages are referenced to Benchmark 0 (BM0), located in a gentle transition from low to high marsh at an elevation of about 650 cm above mean sea level. To decimetre precision, BM0 corresponds to average bankfull level for the purpose of distinguishing below-marsh and over-marsh tides. Irregular terrain and slumping along the creek, however, make a more precise definition of mean bankfull level difficult. Stage was continuously recorded to a vertical precision of $\pm 1{\cdot}6$ cm and at a resolution of 15 min.

During summer 1994, depth-integrated suspended sediment concentrations at the creek mouth did not exceed 10 mg l^{-1}, virtually all of which was in the silt/clay range, and there was little evidence of accumulation of fines over sand deposits. To estimate net bedload transport directions and rates, the migration of several sand dunes on the creek bed was monitored over three days of moderate tides (high tide between +4 cm and +66 cm with respect to BM0). This method does not account for silts, clays or fine sands travelling in suspension or involved in marsh surface accretion, but it reveals the general trend of bedload transport, the main factor in the evolution of creek morphology. The dimensions, migration direction and distance travelled over two tidal cycles were measured for 19 test dunes on ten sand deposits on the creek bed (Figure 10). Net sand transport in kilograms per second per metre width was estimated using the formula

$$q_s = 0{\cdot}5(c_s H D_s)$$

where c_s is the dune migration rate (m s^{-1}), H is the dune height (m) and D_s is the bulk density of sand deposits (1600 kg m^{-3}). This equation assumes each dune to have a triangular profile (Simons and Sentürk, 1992). Because tidal current data were collected away from the channel margin bars, where most bed material transport occurred and dunes were observed, available flow records were unsuitable to model local bed material transport rates given the complex temporal and spatial variability of flow forces in estuarine channels.

RESULTS

Creek morphology and sedimentology

Lower Dipper Harbour Creek can be conveniently broken into three major morphological reaches, along with a short fourth reach near the upvalley end of the marsh (Figure 1). This division is based on channel shape and sedimentology.

Reach 1 begins at the creek mouth and extends over two broad meanders to a large island at cross-section T8 (about 790 m from the mouth; see Figure 1). The channel in Reach 1 is wide, peaking at 100 m over the flood delta gravel bar just upvalley of the bridge and averaging 25 m for the rest of the reach. The bed material is sandy gravel (D_{50} of 1·7 mm at B1, over 3·35 mm at B2), gradually becoming finer toward the upvalley end of the reach (Figures 1 and 2). The long profile in Reach 1 shows a rapid rise in thalweg elevation from the mouth to cross-section T4, and then a gentler rise until T8 (Figure 2).

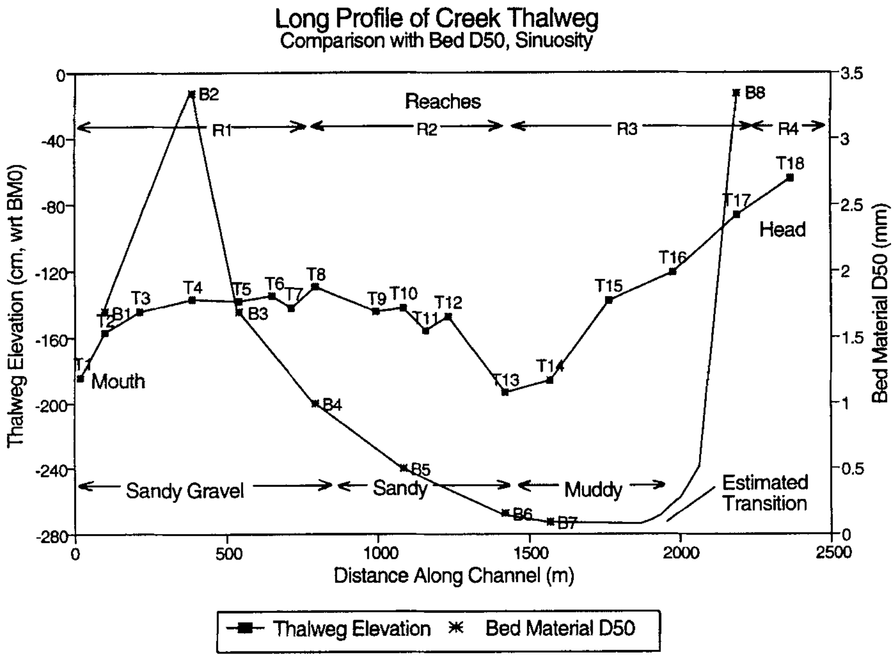

Figure 2. Upvalley changes in thalweg elevation and median size of bed materials. Height datum (BM0) and chatacteristics of Reaches 1, 2, 3 and 4 are discussed in the text

The large island at T8 marks the beginning of Reach 2, which is considerably more sinuous than Reach 1 (sinuosity is 2·8 in Reach 2, 1·6 in Reach 1), with smaller and tighter meanders. The creek rapidly narrows through this region, from 49 m at T8 to 19·5 m at T9, and down to only 6·5 m at T13 (Figure 1). The bed material of Reach 2 is mostly medium to coarse sand (Figure 2), and each meander has a sandy point bar bearing dunes. A striking aspect of this reach lies in the long profile of its thalweg. Instead of a steady upvalley rise, the bed in this reach drops about 50 cm from T8 to its lowest point T13. The lower 1·5 km of the creek's long profile (Reaches 1 and 2) thus has the appearance of a broad hump (Figure 2). In addition to the island at T8, a relic island with a partly infilled backchannel is found between cross-sections T10 and T11. In terms of sinuosity, secondary channels, bedforms and long profile, Reach 2 is morphologically the most complex area of Dipper Harbour Creek.

Reach 3 starts around 1500 m from the creek mouth, and it is the longest and straightest reach (Figure 1). The channel here is quite narrow, its width averaging about 4 m, and the bed is of mud and fine sand with gravel particles in the thalweg. There is no notable pure sand accumulation in this reach, and as Figure 2 shows, the upvalley end of medium sand deposits coincides closely with the upvalley end of the long profile hump at the head of Reach 2. A notable increase in D_{50} around km 1·9 (Figures 1 and 2) reflects pebble size material found in a crude series of pools and riffles downstream of a small freshwater tributary. The creek bed in Reach 3 rises steadily toward the upvalley end of the marsh (Figure 2).

Reach 4, beginning between T17 and T18 (about 2175 m from the mouth; see Figure 1), features very small and tight meanders. Channel width stays around 4 m, and bed material is mostly muddy, though some gravel and rough cobble is found near tributary brooks and along the valley wall. The long profile continues to rise until the end of the tidal portion of the creek. The marsh plants along the creek have gradually become mostly freshwater species by Reach 4. Data on current velocity and bed material transport patterns help explain these morphological and sedimentological observations.

Current velocity asymmetry

Dipper Harbour Creek shows distinct and consistent asymmetry between its flood and ebb flows. As

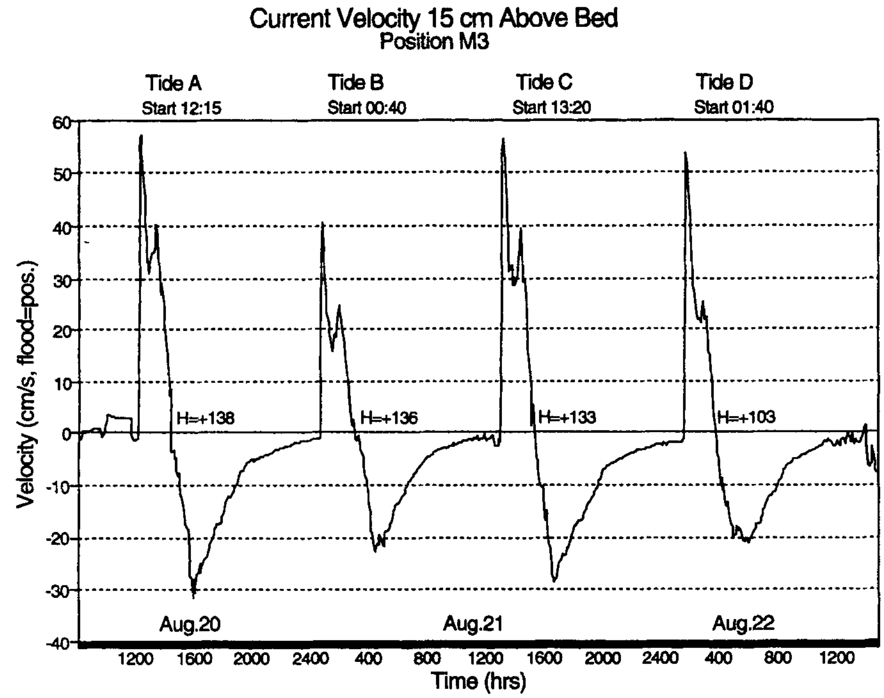

Figure 3. Time series of tidal velocities at 15 cm above the bed for Position M3 during 20–22 August 1993. Height of each tide (H) is given with respect to BM0, corresponding approximately to bankfull stage in the lower creek

Figures 3 to 6 show, the creek's flow regime is strikingly flood-dominant, with peak flood velocities up to three times faster than those of the ebb. For instance, on a spring tide (Figure 3, Tide A with high tide stage $H = +138$ cm with respect to Benchmark 0) on 20 August 1993, the peak recorded flood velocity 15 cm above the bed at Position M3 (at the head of Reach 1; see Figure 1) was 57 cm s^{-1}, compared to a peak

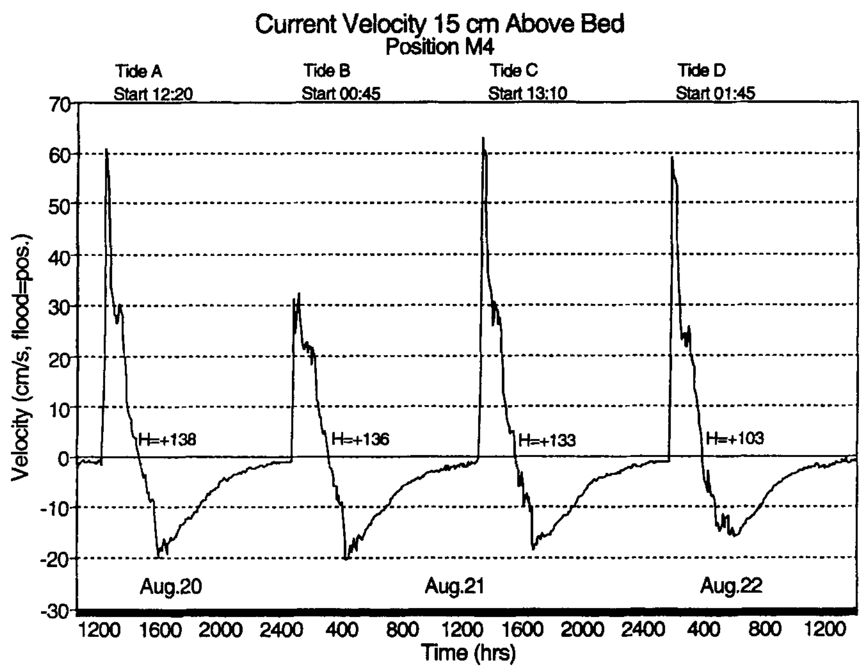

Figure 4. Time series of tidal velocities at 15 cm abive the bed for Position M4 during 20–22 August 1993. Height of each tide (H) is given with respect to BM0

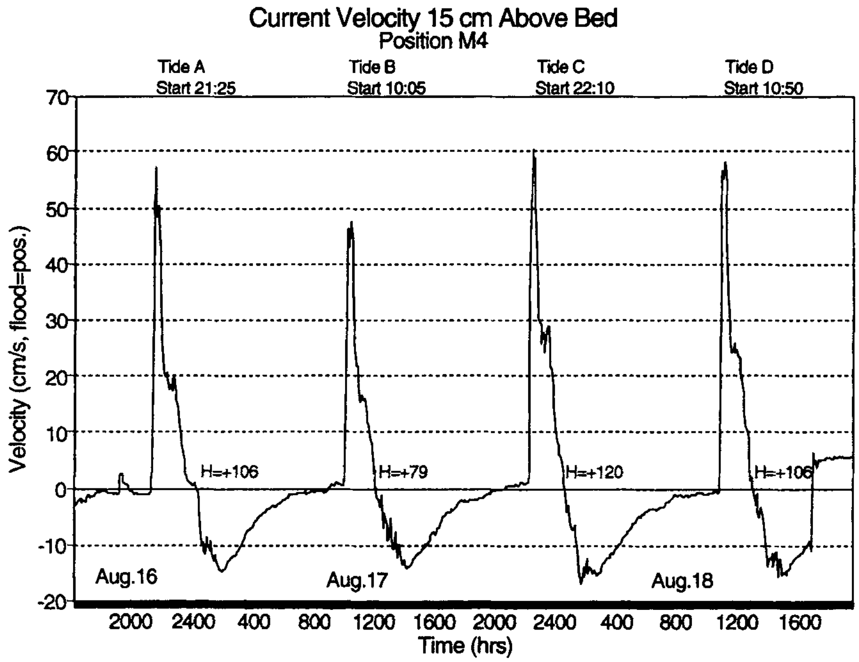

Figure 5. Time series of tidal velocities at 15 cm above the bed for Position M4 during 16–18 August 1993. Height of each tide (H) is given with respect to BM0

of 32 cm s^{-1} on the ebb (for further examples, see also Figure 3, Tides C and D; Figures 4, Tides A, C and D). The greatest recorded asymmetries occurred at Position M4 during a set of moderate tides between 16 and 17 August (Figure 5). On 17 August (Tide C, H = +120 cm), a maximum flood velocity of 60 cm s^{-1} and a maximum ebb of only 17 cm s^{-1} were observed. The pattern of strong, consistent flood dominance also holds

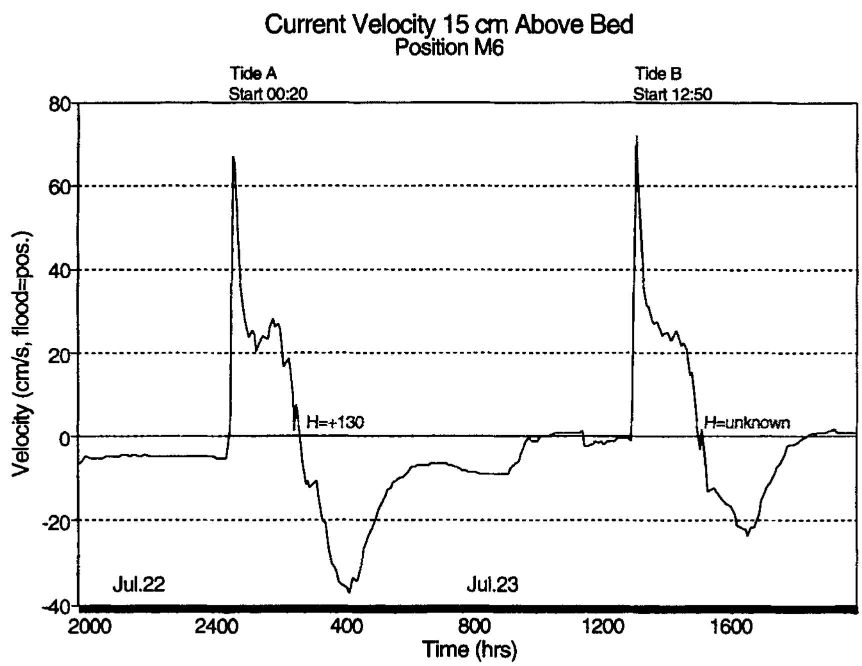

Figure 6. Time series of tidal velocities at 15 cm above the bed for Position M6 during 22–23 July 1993. Height of each tide (H) is given with respect to BM0

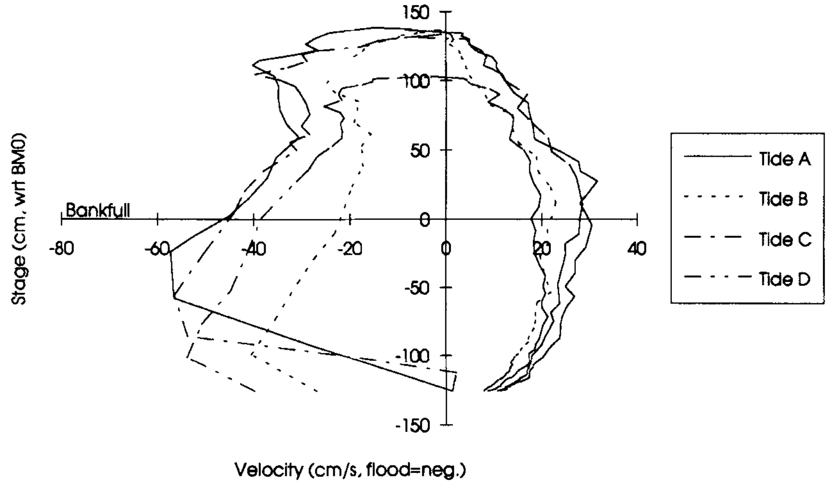

Figure 7. Tidal stage–velocity curves for position M3, 20–22 August 1993. The elevation of BM0 corresponds to the lower edge of the high marsh

farther upvalley. At Position M6, the farthest upvalley current metering station (in Reach 3 at 1680 m from the mouth), maximum flood flow on a spring tide on 23 July (Figure 6, Tide A, H = +130 cm) was 67 cm s^{-1}, compared to a peak ebb of 36 cm s^{-1}, and on the next tide (Figure 6, Tide B) the peak flood velocity was 72 cm s^{-1} while ebb flow peaked at 24 cm s^{-1}.

Recorded data are insufficient to resolve precise upvalley trends in peak tidal velocity, since the only simultaneous data covering full tidal cycles come from Positions M3 and M4. These locations were quite near one another, and a total of only four tides of simultaneous data were collected. The data do, however, argue

Table I. Timing of peak flow velocities for each tide cycle presented in Figures 4–7. Peak flood lag is time between onset of flood and peak flood velocity; peak flood lead is time between peak flood velocity and high tide; peak ebb lag is time between high tide and peak ebb velocity. This sample includes a range of tides from average to spring conditions

Deployment	Tide	Time of start of flood	Peak flood lag (min)	Peak flood lead (min)	Peak ebb lag (min)	Flood duration (min)	Ebb duration (min)	High water (cm, wrt BM0)
20–22 August	A	12:15	15	125	85	140	605	+138
Position M3	B	00:40	10	135	85	145	600	+136
	C	13:05	15	130	80	145	610	+133
	D	01:40	10	125	130	135	–	+103
20–22 August	A	12:20	10	135	75	145	600	+138
Position M4	B	00:45	25	125	75	150	595	+136
	C	13:10	10	140	70	150	605	+133
	D	01:45	10	125	125	135	–	+103
16–18 August	A	21:25	10	170	95	180	580	+106
Position M4	B	10:05	25	105	120	130	595	+79
	C	22:10	15	135	110	150	610	+120
	D	10:50	20	120	90	140	–	+160
22–23 July	A	00:20	10	130	90	140	610	+130
Position M6	B	12:50	15	120	90	135	–	?
Averages			14	130	94	144	601	120
Standard deviations			5·5	14·4	19·4	12·1	9·4	18·8

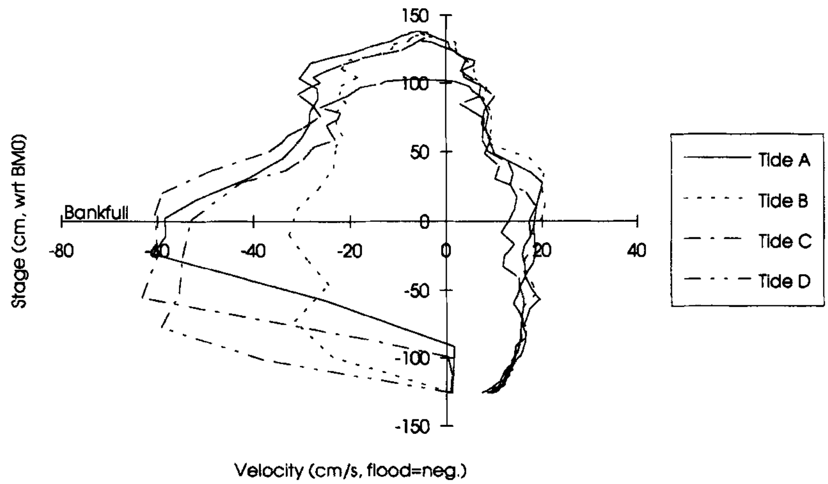

Figure 8. Tidal stage–velocity curves for position M4, 20–22 August 1993

against any clear lowering of peak flow strengths at least as far upvalley as Position M6, despite the observed upvalley fining of bed material. For example, the peak flood velocities on a high spring tide at Positions M3 and M4 (Figures 3 and 4, Tide A, H= +138 cm) were exceeded by the velocities on a somewhat smaller tide farther upvalley at Position M6 (Figure 6, Tide A, H= +130 cm).

Timing and stage of peak flows

The precise timing of maximum flows within the tidal cycle is quite regular over a sample of moderate to strong tides in Dipper Harbour Creek (Table I). Peak flood flow always occurs early in the tide, on average about 15 min after the tidal wave enters the creek and some 2 h 10 min before high slack water. Peak ebb

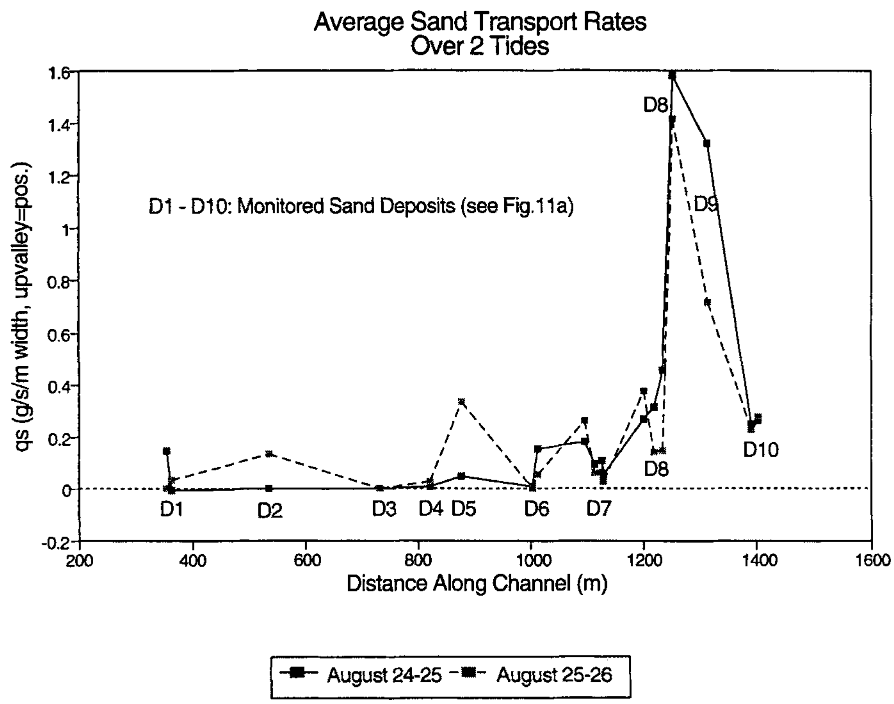

Figure 9. Patterns of net sand discharge associated with dune migration during 24–26 August 1993

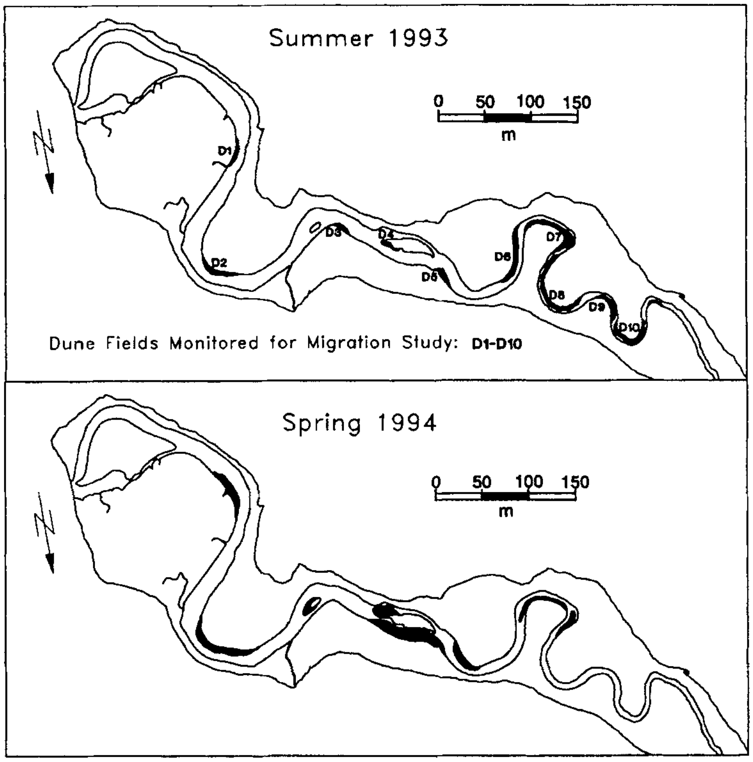

Figure 10. Location map: major sand deposits in summer 1993, spring 1994. Locations of sand deposits monitored for dune migration during summer 1993 are also shown

velocity appears somewhat less consistent in time, perhaps because it is seldom much stronger than the average ebb flow. The lag time from high tide to peak ebb velocity varies between 75 and 130 min, averaging about 95. The implication of this timing, as illustrated in the stage–velocity curves of Figures 7 and 8, is that the peak flood velocities occurs at low stages, generally between 0·5 and 1 m below the average bankfull level of BM0. Peak ebb flow generally occurs quite near bankfull level, a stage at which the high marsh is draining into the creek.

A notable feature of Dipper Harbour Creek's flow regime is a brief velocity surge consistently occurring at high stages during the flood phase of many tides (Figures 3–6) and lasting an average of 20 min. Such surges have been documented in other tidal creeks (Pethick, 1980; Bayliss-Smith *et al.* 1979; French and Stoddart, 1992), where they often constitute the flood velocity peak. In Dipper Harbour Creek, however, observed flood surges were always dwarfed by the earlier flood velocity maximum, and sometimes were so small as to be nearly indistinguishable. A good illustration of a typical flood surge for Dipper Harbour Creek is provided by the data from the afternoon of 20 August at Position M3 (Figure 7, Tide A). After a flood peak of $57\,\mathrm{cm\,s^{-1}}$ at stage -58 cm relative to BM0, the flow gradually slowed to $31\,\mathrm{cm\,s^{-1}}$ at stage $+60$ cm before surging back up to $40\,\mathrm{cm\,s^{-1}}$ at stage $+111$ cm and then steadily declining to zero at high slack water. A similar pattern was simultaneously recorded at Position M4 (Figure 8, Tide A). The high-stage surge is quite stable in its relative prominence over a variety of over-marsh tides, as shown in the two velocity–time plots from Position M4 (Figures 4 and 5). Surges can last between 5 and 45 min and are generally stronger than the peak ebb velocity, increasing the relative effectiveness of flood flows for sediment transport in this system.

Sand transport patterns

Data on the migration of sand dunes, the most common bedforms in Dipper Harbour Creek, provide a direct indication of net bed material transport in the channel during summer flow conditions. Given the

difficulties in empirically modelling bed materials transport rates over gravelly sand channel beds on the basis of sparse velocity data (predictions made particularly uncertain because of the hysteresis in turbulence intensity in tidal flows (Gordon, 1975)), the actual observations of bed material transport rates presented here are especially valuable. Local near-bed sand discharge rates obtained in late August 1993 are presented as a function of upchannel distance in Figure 9. Few sand accumulations were found in Reach 1 (Figure 10). These dunes show slight upvalley migration but contribute little sand discharge, due mostly to their small amplitude ($\leqslant 10$ cm). Sand discharge rises moderately in the first meander of Reach 2. Deposits in this area are extensive but the dunes are still small and show moderate upvalley migration, resulting in mean sand discharges of less than $0{\cdot}4\,g\,s^{-1}$ per metre width.

The region of fastest dune migration, with dune crest advance rates of up to 0·5 m per tide, is at deposit D8 on a point bar in Reach 2's second meander (between cross-sections T11 and T12). The dunes here are bigger (height between 4 cm and 27 cm), the largest being permanently submerged, and all show rapid upvalley migration, causing a sudden jump in sand discharge (up to nearly $1{\cdot}6\,g\,s^{-1}\,m^{-1}$). Past this bend, sand deposits remain extensive until the end of Reach 2 (about 1500 m from the mouth), but dune heights drop sharply, producing lower sand discharge (between 0·2 and $0{\cdot}3\,g\,s^{-1}\,m^{-1}$). Overall, the dominant direction of near-bed sand transport is upvalley, with no significant downvalley dune migration observed.

Though dune migration experiments were not replicated, a major shift in the location of the creek's sand deposits was observed in May 1994, just after the spring snowmelt freshet (Figure 10). The main zone of sand accumulation had shifted downvalley from the upvalley end of Reach 2 to the region around the two islands near the head of Reach 1. With the spring 1994 increase in medium sands on the bed in this area around the islands, the median bed material caliber dropped from 1 mm in summer 1993 to 0·5 mm in spring 1994 (site B4; see Figure 1). Within Reach 2, many of 1993's most prominent dune fields were gone or much reduced in spring 1994. The flood-aligned point bar at D5 just upvalley of the large island near T8 had expanded and become ebb-aligned in 1994 (Figure 10). The overall effect of the spring flood was thus a downvalley shift in the creek bed's sand and a general change among the sand deposits from flood alignment to ebb alignment.

DISCUSSION

Tidal asymmetry

Many documented tidal creek systems are ebb-dominant (Pestrong, 1965; Bayliss-Smith *et al.*, 1979; Reed *et al.*, 1985; French and Stoddart, 1992), or only weakly flood-dominant (Myrick and Leopold, 1963; Pethick, 1980). Bayliss-Smith *et al.*, (1979) present stage–velocity curves for Lady Creek on England's Norfolk Coast (their figure 4) which resemble those for the present study, although they claim the pattern of dominant low-stage peak flood velocity is restricted to below-marsh tides incapable of producing the usual high-stage flood surge. In Dipper Harbour Creek, the dominant early flood surge is the normal pattern on observed 1993 spring tides, some nearly 1·5 m overbank, and precedes a secondary high-stage flood surge.

French and Stoddart (1992) mention an early flood velocity pulse caused by flow over a sand bar across the mouth of East Creek, also on the Norfolk Coast. In Dipper Harbour Creek, the predominance of the early flood peak may partly be caused by a discontinuity in the relation between stage and tidal prism, tied to the hump observed in the lower creek's long profile. Once the flood stage exceeds approximately −130 cm (wrt BM0), water begins to fill the long 'trough' in the creek bed upvalley of cross-section T8 (Figure 2). Based on measured channel dimensions, approximately $1000\,m^3$ of flood water is required to fill this trough up to the −130 cm stage. Observed low-stage surges begin at a stage between −100 cm and −50 cm with respect to Benchmark 0 (i.e. a water depth of between 30 cm and 80 cm over the thalweg at T8) (Figures 2, 7 and 8). Given the small wetted cross-section at these early stages, high flood velocities may develop to fill in the trough in the long profile between kms 0·8 and 1·6 (Figure 2). As suggested by Boon (1975) and Pethick (1980), the asymmetric form of the tidal wave as it enters the creek may also play a role in the flood-dominance of Dipper Harbour Creek. The shoaling Fundy tidal wave's steep flood limb produces a shorter flood and a more sudden change in stage on the flood than on the gentler ebb limb, necessitating faster flood velocities for the full tidal prism to enter the system.

While the early flood peak, which is not uncommon in macrotidal estuaries (Wright *et al.*, 1975; Dalrymple and Zaitlin, 1989) may be reinforced here by Dipper Harbour Creek's distinctive long profile, the secondary high-stage flood surge and peak ebb flow are similar to those found in other creek systems. The former, occurring about 1 m above bankfull, may stem from the sudden increase in tidal prism as the high marsh floods, producing a surge in velocities in the creek (Bayliss-Smith *et al.*, 1979; French and Stoddart, 1992). The ebb peak, for its part, occurs closer to bankfull stage, and may be produced by water inputs from gravity drainage of the marsh surface (French and Stoddart, 1992).

Sedimentological patterns and the balance between fluvial and tidal forces

Dalrymple *et al.* (1992) predict a macrotidal estuary's sinuous central reach to contain the balance point of fluvial and tidal energy. On a yearly basis, this appears to be the case for Dipper Harbour Creek.

On most of the over-marsh tides monitored, peak flood velocity at 15 cm above the bed exceeded peak ebb velocity by over $20\,cm\,s^{-1}$ at the same measuring station. Given constant measuring height and local bed roughness, the much stronger flood velocities indicate significantly higher bed shear stresses on the flood. This higher concentration of flood energy appears to translate into higher capacity for sand transport on the flood than on the ebb, despite the latter's longer duration. Indeed, the data on sand dune migration in Dipper Harbour Creek point to predominant upvalley sediment transport, with very little bed-calibre sand leaving the creek during the summer study period. Yet in summer 1993, near cross-section T13 at the upvalley end of Reach 2, a sharp transition existed beyond which medium sand-class bed material disappeared (Figure 2). Considering the absence of any clear upvalley damping of flood strength at least up to Position M6 in Reach 3 and the rapid advance rates of some sand dunes, some alternative mechanism is required to explain how sand bedforms have been kept from migrating upvalley of Reach 2 over time.

The spring snowmelt freshet from Dipper Harbour Creek's basin seems to provide the required flushing mechanism to check the upvalley movement of sand. The timing of the ice breakup and freshet peak is highly unpredictable in small maritime basins in Canada and could not be directly observed in 1994, but the magnitude of spring 1994 runoff was not unusual for this area. Based on the channel morphology at the head of the tidal creek (Ayles, 1993) the spring flood yields a bankfull flow of approximately $2\text{–}3\,m^3\,s^{-1}$ and should produce freshwater flow entering the tidal reaches at an average of approximately $0{\cdot}8\,m\,s^{-1}$ in the upper marsh. Though the semidiurnal flood flow would still affect the creek during this period, it would be slowed somewhat by the freshet, and more importantly, the longer-duration ebbward flow might be amplified enough to establish ebb-dominant sand transport during the brief snowmelt season.

Visual evidence after the freshet in spring 1994 strongly suggests that such a mechanism was at work in Dipper Harbour Creek that year (Figure 10). The fresh accretion of major sand deposits around and downvalley of the two islands (at T6, T8; see Figure 1), the reduced area of sand deposits in Reach 2 and the ebb-alignment of most sand formations all points to a general downvalley shift of the main zone of sand accumulation after the 1994 freshet. This is further supported by the observed fining of bed material near the large island at T8 between August 1993 and May 1994, indicating a relative local increase in sand-class material. Although the precise timing of breakup is unpredictable, snowmelt high flows are a regular, dominant feature of coastal hydrology in this region. Thus, the 1993–94 data strongly suggests the existence of a seasonal cycling of sediment within Dipper Harbour Creek, with a body of mostly sandy bedload migrating upvalley under tidal forces over most of the year and then being pushed back downvalley by the spring meltwater freshet. Multi-year observations are required to confirm the scope and variability of this cycling effect.

The summer location of the seasonally shifting sand body appears to be centred on Reach 2, the upstream limit of which coincided with the low point in the thalweg long profile in summer 1993 (Figure 2). Similar to what Wright *et al.* (1975) suggested for the Ord River estuary, the accumulation of these sands in Reach 2 may be responsible for Dipper Harbour Creek's greater morphological complexity between cross-sections T6 and T13, as mid-channel bars and point bars built of mobile sands are important elements in the development of the islands and tight meanders characteristic of this reach. The summertime upvalley migration of sands in Reach 2 conflicts somewhat with the mud-dominated sinuous central reach predicted by the model of Dalrymple *et al.* for estuaries with more sustained freshwater input. Still, on a yearly basis, the observed

counteraction of spring and summer bed sand transport in Reach 2 suggests that this highly sinuous central reach is indeed at the average balance point between tidal and fluvial forces within Dipper Harbour Creek. This conclusion agrees with similar finding from the Ord (Wright *et al.*, 1975) and Salmon River (Dalrymple and Zaitlin, 1989) estuaries.

CONCLUSION

During the study period of summer 1993, Dipper Harbour Creek's hydraulic regime resembled that of a macrotidal salt marsh channel, with a geomorphically active tidal flow. This flow featured the reversals and velocity surges characteristic of tidal creeks, but was flood-dominant with an exceptionally strong, consistent peak in flow energy early in the flood. Dipper Harbour Creek departs from true tidal channels, however, in its yearly burst of fresh meltwater discharge and the consequent disruption of purely tidal sediment transport. The seasonal cycling of sand observed on the creek bed, which results from the interaction of these two flow regimes, is an expression of the system's true estuarine character.

Dipper Harbour Creek is morphologically similar to the idealized tidally dominated estuary in Dalrymple *et al.*'s (1992) classification, though there are some hydrodynamic and sedimentological differences. These stem from the middle reach, where, on a yearly basis, tidal and fluvial forces achieve balance, and which displays the best developed bars and sinuosity. While Dalrymple *et al.* predict this reach to contain the estuary's finest sediments, by mid- to late summer in Dipper Harbour Creek it is an accumulation zone for mobile sands, with finer muds found higher upvalley. Thus, Dipper Harbour Creek is a two-faceted system: in the short term, it acts as a flood-dominant tidal creek, and in the long term it can be classified as a particular example of tide-dominance among the estuaries of Dalrymple *et al.* Further research is necessary to ascertain the extent to which the patterns observed in Dipper Harbour Creek apply to other small macrotidal estuaries.

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